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Erosion-driven uplift in the Gamburtsev Subglacial Mountains of East Antarctica

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Abstract

The relative roles of climate and tectonics in mountain building have been widely debated. Central to this debate is the process of flexural uplift in response to valley incision. Here we quantify this process in the Gamburtsev Subglacial Mountains, a paradoxical tectonic feature in cratonic East Antarctica. Previous studies indicate that rifting and strike-slip tectonics may have provided a key trigger for the initial uplift of the Gamburtsevs, but the contribution of more recent valley incision remains to be quantified. Inverse spectral (free-air admittance and Bouguer coherence) methods indicate that, unusually for continents, the coherence between free-air gravity anomalies and bedrock topography is high (>0.5) and that the elastic thickness of the lithosphere is anomalously low (<15 km), in contrast to previously reported values of up to ~ 70 km. The isostatic effects of two different styles of erosion are quantified: dendritic fluvial incision overprinted by

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Alpine-style glacial erosion in the Gamburtsevs and outlet glacier-type selective linear erosion in the Lambert Rift, part of the East Antarctic Rift System. 3D flexural models indicate that valley incision has contributed ca. 500 m of peak uplift in the Gamburtsevs and up to 1.2 km in the Lambert Rift, which is consistent with the present-day elevation of Oligocene–Miocene glaciomarine sediments. Overall, we find that 17–25% of Gamburtsev peak uplift can be explained by erosional unloading. These relatively low values are typical of temperate mountain ranges, suggesting that most of the valley incision in the Gamburtsevs occurred prior to widespread glaciation at 34 Ma. The pre-incision topography of the Gamburtsevs lies at 2–2.5 km above sea-level, confirming that they were a key inception point for the development of the East Antarctic Ice Sheet. Tectonic and/or dynamic processes were therefore responsible for ca. 80% of the elevation of the modern Gamburtsev Subglacial Mountains.

Keywords:

East Antarctica, gravitational admittance, flexure, erosion, landscape evolution, paleotopography

1 **1. Introduction**

2 The Gamburtsev Subglacial Mountains (GSM) are located beneath Dome A
3 of the East Antarctic Ice Sheet (EAIS) (Fig. 1). Although the GSM cannot be
4 directly observed, the subglacial landscape has recently been revealed by Antarc-
5 tica’s Gamburtsev Province (AGAP) radar, aerogravity and aeromagnetic data,
6 collected during the International Polar Year (2008–2009) (Bell et al., 2011). The
7 GSM exhibit 2–3 km of relief and a landscape heavily dissected by fluvial and
8 glacial valleys that resembles the European Alps (Bo et al., 2009; Creyts et al.,

9 2014; Rose et al., 2013). Flanking the Gamburtsevs are a series of north–south
10 trending basins interpreted as comprising the East Antarctic Rift System (EARS)
11 (Ferraccioli et al., 2011). When compared to other mountain ranges, the Alpine-
12 style geomorphology of the GSM (Creyts et al., 2014) is paradoxical, since they
13 are located atop Precambrian cratonic lithosphere (Heeszel et al., 2013). This
14 problem is compounded because no in situ geological samples from the GSM
15 exist; their lithology, age and structure remain unknown.

16 Unravelling the enigmatic topographic evolution of the GSM is particularly
17 important, because (1) this mountain range is thought to have provided a key nu-
18 cleation site for the development of the EAIS at the Eocene–Oligocene Boundary
19 (DeConto and Pollard, 2003; Rose et al., 2013) and (2) the processes that build
20 intraplate mountains remain poorly understood, and the Gamburtsevs are the most
21 enigmatic intraplate mountain range on Earth.

22 Permian rifting and Cretaceous strike-slip faulting have been advanced as tec-
23 tonic triggers for GSM uplift (Ferraccioli et al., 2011). However, the isostatic
24 response to fluvial/glacial valley incision has been suggested to be responsible for
25 the modern relief and geomorphology of the GSM (Ferraccioli et al., 2011), as has
26 been demonstrated in other mountain ranges (e.g. Champagnac et al., 2007). This
27 isostatic uplift has been quantified using simple 2D flexural models (Ferraccioli
28 et al., 2011), but the 3D distribution of erosion and flexure, as well as the influence
29 of the neighbouring Lambert Rift, have not previously been considered. The aim
30 of this study is to quantify the spatial distribution of Cenozoic fluvial and glacial
31 erosion and the associated isostatic response prior to and during the early stages
32 of EAIS development in order to determine whether this effect was sufficient to
33 drive a substantial part of the uplift of the GSM.

34 To address this question, the AGAP radar and aerogravity data were used to
35 estimate the effective elastic thickness of the lithosphere (T_e) and the amount and
36 distribution of eroded material in the Gamburtsev region. 3D flexural models were
37 used to calculate the resulting flexural uplift induced by valley incision for differ-
38 ent T_e scenarios, and thereby estimate the pre-incision elevation of the GSM. The
39 age of fluvial incision in the GSM was constrained using a landscape evolution
40 model. The main findings are that the processes of valley incision in the GSM
41 predominantly occurred in a temperate climate, and that the Gamburtsevs were at
42 2–2.5 km elevation prior to the Eocene–Oligocene Boundary.

43 **2. Aerogeophysical Data Acquisition and Reduction**

44 The acquisition of AGAP airborne geophysical data took place between 2nd
45 December 2008 and 16th January 2009. Two de Havilland Canada Twin Otter air-
46 craft successfully obtained 120,000 line-km of radio-echo sounding (RES), aero-
47 magnetic and aerogravity data over the GSM and adjacent Lambert Rift. The
48 survey comprised flight lines oriented north–south, with 5 km horizontal spacing.
49 East–west tie lines intersected the main lines every 33 km.

50 *2.1. Surface and Bedrock Topography*

51 Mapping of surface and bedrock topography was carried out using a wing-
52 mounted RES system. RES data were acquired using ice-penetrating radars with
53 a 150 MHz carrier frequency and 15–20 MHz bandwidths, which sample the ice at
54 2 m intervals along the flight-track (Creyts et al., 2014). Kinematic GPS provided
55 location and altitude data accurate to ~5 cm.

56 The two-way travel time (TWTT) for the ice surface reflector was multiplied
57 by the radar velocity in air (300 m/ μ s) to give the terrain clearance of the aircraft.

58 The difference between the altitude of the aircraft and the terrain clearance is
59 the surface elevation. The difference in TWTT between the bed and ice surface
60 reflectors gives the TWTT in the ice, which is depth converted to an ice thickness
61 using an ice radar velocity of $168 \text{ m}/\mu\text{s}$, with an additional 10 m correction for
62 the firn layer. The difference between the surface elevation and the ice thickness
63 gives the bed elevation. Bed elevations were measured relative to the WGS-84
64 ellipsoid. The root mean square (RMS) cross-over error was 64 m (Creys et al.,
65 2014).

66 The radar data were gridded using a ‘nearest neighbour’ gridding routine
67 (GMT’s *nearneighbor* module (Wessel et al., 2013)) with a grid spacing of 1
68 km and search radius of 5 km. To form a complete bedrock topography grid
69 for the East Antarctica, data gaps in the grid were filled using the Bedmap2
70 compilation (Fretwell et al., 2013). This maintained the high resolution of the
71 AGAP data while avoiding excessive computational demand. Grid profiles com-
72 pare favourably with real RES data (Fig. 2). While gridding causes some of the
73 resolution to be lost, the grid picks out the sharp and high local relief observed
74 in the radar data. Radar-derived bedrock topography data are essential for the
75 spectral analysis carried out in this study, as they guarantee independence of the
76 gravity and topography grids.

77 2.2. Aerogravity

78 The UK aircraft acquired aerogravity data using a LaCoste-Romberg S-83 air-
79 sea gravimeter (Jordan et al., 2007). The lines were flown in a stepped pattern with
80 a maximum altitude of 4,600 m over Dome A. The US aircraft used a Sander Geo-
81 physics AIRGrav airborne gravity system (Studinger et al., 2008); these lines were
82 flown at a constant terrain clearance not in excess of 500 m. Corrections were ap-

83 plied for the vertical accelerations that act on the aircraft, the Eötvös effect (which
84 depends on speed and heading), and the ‘cross-coupling’ between the horizontal
85 and vertical accelerations. Data were tied to a base station at McMurdo Station
86 using a LaCoste-Romberg land gravimeter, thereby converting relative gravity to
87 absolute values. Gravity data from the two aircraft were combined and filtered us-
88 ing a 9 km half-wavelength low-pass space-domain kernel filter (Holt et al., 2006).
89 They were then upward continued to a uniform altitude of 4,600 m above the el-
90 lipipsoid (corresponding to the maximum flight altitude). After reduction, filtering
91 and upward continuation, the overall RMS cross-over error of the free-air gravity
92 anomaly (FAA) data was 2 mGal.

93 The FAA data were ‘nearest neighbour’ gridded with a horizontal spacing of
94 1 km and search radius of 5 km. Long-wavelength Gravity field and steady-state
95 Ocean Circulation Explorer (GOCE) satellite gravity models (Yi et al., 2013) were
96 resampled to 1 km, upward continued to the 4,600 m geodetic datum and used to
97 fill in data gaps surrounding the main AGAP FAA grid. This formed a complete
98 FAA grid for East Antarctica (Fig. 2).

99 Gravity anomalies arise from undulating interfaces across which there is a
100 density change. In order to calculate a Bouguer correction, the gravity effects of
101 (1) the ice surface and (2) the ice-bed interface were calculated using Parker’s
102 expression for the gravity effect of an undulating interface of uniform density
103 contrast (Parker, 1972) (Supplementary Fig. 1). The applied reduction densities
104 for air, ice and rock were 0, 915 and 2670 kgm⁻³, respectively. The correction for
105 the ice surface was subtracted from the FAA prior to spectral analysis. Subtraction
106 of both corrections from the FAA produced the complete Bouguer anomaly (Fig.
107 2), which was median filtered to remove wavelengths shorter than 18 km to match

108 the filtered FAA.

109 **3. Methods**

110 *3.1. Gravitational Admittance and Coherence*

111 *3.1.1. Theory*

112 There are two standard inverse (spectral) methods used to estimate the effective elastic thickness of the lithosphere, T_e , using gravity and topography data: 113 the free-air admittance and the Bouguer coherence. The admittance, $\mathbf{Z}(k)$, is the 114 linear transfer function between the gravity anomaly and topography in the frequency domain (e.g. Kirby, 2014). 115

$$\mathbf{Z}(k) = \frac{\langle \Delta \mathbf{g}(\mathbf{k}) \cdot \mathbf{H}^*(\mathbf{k}) \rangle}{\langle \mathbf{H}(\mathbf{k}) \cdot \mathbf{H}^*(\mathbf{k}) \rangle} \quad (1)$$

117 $\Delta \mathbf{g}(\mathbf{k})$ is the Fourier transform of the observed gravity anomaly, $\mathbf{H}(\mathbf{k})$ is the 118 Fourier transform of the observed topography, $\mathbf{k} = (k_x, k_y)$ is the 2D wavenumber 119 and $k = |\mathbf{k}|$, $*$ denotes the complex conjugate and $\langle \rangle$ indicates annular averaging of 120 the spectral estimates. Another useful spectral parameter is the coherence, $\gamma^2(k)$, 121 which is expressed as (Kirby, 2014)

$$\gamma^2(k) = \frac{|\langle \Delta \mathbf{g}(\mathbf{k}) \cdot \mathbf{H}^*(\mathbf{k}) \rangle|^2}{\langle \Delta \mathbf{g}(\mathbf{k}) \cdot \Delta \mathbf{g}^*(\mathbf{k}) \rangle \langle \mathbf{H}(\mathbf{k}) \cdot \mathbf{H}^*(\mathbf{k}) \rangle} \quad (2)$$

122 The coherence is essentially the square of the Pearson product-moment correlation 123 coefficient between gravity and topography computed in the frequency domain 124 (Kirby, 2014). A high coherence indicates that a large fraction of the gravity 125 anomaly is caused by the topography. In this study, 0.5 is used as the threshold 126 between high and low coherence. The phase of the admittance, $\phi(k)$, is defined by 127 (Watts, 2001)

$$e^{-i 2 \phi(k)} = \frac{\mathbf{Z}(k)}{\mathbf{Z}^*(k)} \quad (3)$$

128 Where the coherence is high, the phase of the admittance should be close to zero.

129 The bedrock topography, free-air and Bouguer anomaly grids were projected
130 into a customised Lambert conformal conic projection (with central meridian
131 80.0°E; southern and northern parallels 83.0°S and 77.0°S; and central scale fac-
132 tor 1:1) in order to minimise distortion. The admittance and coherence were cal-
133 culated using a standard multitaper method (following McKenzie and Fairhead,
134 1997; McKenzie, 2003; Pérez-Gussinyé et al., 2004). The calculation was carried
135 out for a particular window in the gravity and topography grids. Too small a win-
136 dow will truncate the long wavelengths that characterise high T_e s, causing a bias
137 towards low values. Too large a window will incorporate different geological fea-
138 tures; the recovered T_e will be ambiguous. The calculation was therefore carried
139 out for four grid windows of increasing size centred on the GSM (Fig. 3).

140 3.1.2. Elastic Plate Modelling

141 Within the range of wavelengths over which elastic flexure of the lithosphere
142 is important (100–1000 km) where the ‘roll-over’ from high to low admittance
143 occurs, the shape of the admittance and coherence functions are dependent on the
144 rigidity of the lithosphere. Assuming that the lithosphere behaves as an elastic
145 plate over geological timescales, the calculated free-air admittance was compared
146 to a model admittance for a flexed elastic plate overlying an inviscid fluid. The
147 model assumes that the plate is subject to surface loading only and that the den-
148 sity of the crust is uniform and equal to that of the load (the topography). The
149 theoretical admittance for this model is given by (Watts, 2001)

$$Z(k) = 2 \pi G (\rho_c - \rho_i) e^{-kd} (1 - \Phi_e(k) e^{-kt}) \quad (4)$$

150 where

$$\Phi_e(k) = \left[\frac{D k^4}{(\rho_m - \rho_c) g} + 1 \right]^{-1} \quad (5)$$

151 is the flexural response function, and

$$D = \frac{E T_e^3}{12 (1 - \nu^2)} \quad (6)$$

152 is the flexural rigidity. ρ_c, ρ_i (915 kgm^{-3}) and ρ_m (3330 kgm^{-3}) are the densities of
 153 the topography/crust, ice and mantle respectively; d is the mean distance between
 154 the observation datum (4,600 m above the ellipsoid) and the ice-bedrock interface;
 155 t is the mean crustal thickness; G is the universal gravitational constant; g is the
 156 acceleration due to gravity; E is Young's modulus (100 GPa); and ν is Poisson's
 157 ratio (0.25).

158 At wavelengths shorter than the isostatic rollover ($k \geq 0.15 \text{ radkm}^{-1}$), topog-
 159 raphy is uncompensated and the admittance is given by (Watts, 2001)

$$Z(k) = 2 \pi G (\rho_c - \rho_i) e^{-kd} \quad (7)$$

160 Taking the logarithm of both sides yields

$$\log_{10} Z(k) = -kd \log_{10} e + \log_{10} (2 \pi G (\rho_c - \rho_i)) \quad (8)$$

161 $\log_{10} Z(k)$ was plotted against k and a straight line was fitted to the interval corre-
 162 sponding to the uncompensated topography ($0.15 \leq k \leq k(\gamma^2 = 0.5) \text{ radkm}^{-1}$) by
 163 linear regression (Fig. 3). The interval is capped where the coherence, γ^2 , falls be-
 164 low 0.5, which indicates topography no longer dominates the gravity signal. The
 165 mean ice-bedrock density contrast ($\rho_c - \rho_i$) and depth (d) were determined from
 166 the intercept and the gradient, respectively (Eq (8)). The two remaining free pa-
 167 rameters in the model are t and T_e . Theoretical admittance curves were calculated

168 for a range of t/T_e combinations. The statistical best-fitting combination for each
 169 window is that which minimised the root mean square (RMS) misfit between the
 170 observed and theoretical curves (Fig. 4).

171 Variation in T_e with window size was illustrated by computing the isostatic
 172 response function (IRF) (Watts, 2001),

$$\varphi_e(k) = \frac{Z(k)}{2 \pi G (\rho_c - \rho_i) e^{-kd}} \quad (9)$$

173 which normalises the admittance for d and $\rho_c - \rho_i$ for each window (Fig. 4). The
 174 theoretical IRF for an elastic plate model is given by (Watts, 2001)

$$\varphi_e(k) = 1 - \Phi_e(k) e^{-kt} \quad (10)$$

175 $\varphi_e(k)$ was calculated for a range of T_e values and compared to the observed IRF
 176 (Fig. 4).

177 The coherence between the Bouguer anomaly and bedrock topography - the
 178 ‘Bouguer coherence’ - was also modelled for a flexed elastic plate overlying an in-
 179 viscid fluid. The mean crustal density and thickness for each window derived from
 180 the free-air admittance were used for each Bouguer coherence model. Theoretical
 181 cal Bouguer coherence curves were calculated following the approach of Forsyth
 182 (1985), which incorporates internal (‘buried’) loads with a topographic expres-
 183 sion and assumes that surface and buried loads are incoherent. For each window,
 184 the best-fitting model T_e was that which minimised the RMS misfit between the
 185 observed and theoretical Bouguer coherence (Supplementary Fig. 2).

186 3.2. *Spatial Distribution of Eroded Material*

187 If the spatial distribution of erosion is non-uniform, it is possible for peak
 188 elevations to increase, because local erosion is less than uplift driven by the flex-

189 ural isostatic response to regional erosion. The amplitude and wavelength of the
190 flexural response are dependent on T_e .

191 Quantification of the spatial distribution of eroded material requires the con-
192 struction of a peak/summit accordance surface. This is a 3D surface representing
193 the restoration of eroded material to the topography without accounting for the
194 associated isostatic response (Champagnac et al., 2007). In order to construct the
195 accordance surface, the GSM topography was first adjusted for the removal of the
196 present-day ice load. The method used to compute this adjustment is described in
197 the paragraph at the end of this section. Maximum values in the rebounded topog-
198 raphy grid were isolated using a circular moving window of fixed 15 km radius
199 and designated as peaks (Champagnac et al., 2007) (Fig. 5a). A window radius
200 of 15 km was used to match the approximate wavelength of peaks and valleys in
201 the GSM. It is assumed that these peaks are remnants of a palaeo-surface, and that
202 incision into this surface has not significantly altered the peaks. A preliminary
203 attempt to identify flat-topped surfaces was abandoned, because the GSM are too
204 heavily incised and because the resolution of the topography grid was insufficient
205 to calculate a slope grid and identify flat surfaces.

206 A surface was smoothly interpolated between the peaks using GMT's (Wessel
207 et al., 2013) continuous curvature tensional spline algorithm (with a tension factor
208 of 0.5) and smoothed with a 100 km Gaussian filter. Subtracting the ice-rebounded
209 topography (Fig. 5a) from the peak accordance surface (Fig. 5b) gives a map of
210 eroded material (Fig. 5c). This method of constructing a peak accordance surface
211 assumes that the erosion of the peaks is negligible compared to the erosion in the
212 valleys; the calculated amount of eroded material is a minimum estimate. Because
213 of the inaccessibility of the GSM, there are no constraints on peak erosion from

214 thermochronology, cosmogenic nuclide exposure dating or identification of flat-
 215 topped peaks. Another limitation is the difficulty of fitting a single surface to a
 216 landscape so heavily dissected and affected by multiple regimes of erosion.

217 The flexural response to the removal of the ice sheet and the eroded material
 218 was calculated by solving the general flexure equation for the application of a 2D
 219 (un)load, $h(x, y)$, to an elastic plate overlying an inviscid fluid.

$$\nabla^2 \left[D(x, y) \nabla^2 w(x, y) \right] + (\rho_m - \rho_{infill}) g w(x, y) = (\rho_{load} - \rho_{displace}) g h(x, y) \quad (11)$$

220 The density of the load (ρ_{load}) was assumed to be 915 kgm^{-3} for ice and 2670
 221 kgm^{-3} for eroded bedrock. A mantle density (ρ_m) of 3330 kgm^{-3} was used, and the
 222 material displaced by the (un)loading ($\rho_{displace}$) and infilling the flexure (ρ_{infill}) was
 223 assumed to be air, with a density of 0 kgm^{-3} . The same T_e was used to calculate
 224 the flexure ($w(x, y)$) due to both the ice unloading and erosional unloading. We
 225 computed the flexure for a variety of T_e scenarios based on the results of the
 226 spectral modelling in this study and previously reported estimates (Ferraccioli
 227 et al., 2011) in order to test the sensitivity of the magnitude and pattern of flexural
 228 uplift to the rigidity of the lithosphere.

229 **4. Results**

230 *4.1. Effective Elastic Thickness*

231 By fitting a linear regression line to the portion of the admittance curve cor-
 232 responding to the uncompensated topography (Fig. 3c) and solving Eq (8), an
 233 average shallow bedrock density of $2620\text{--}2750 \text{ kgm}^{-3}$ was recovered across the
 234 different grid windows. A decrease in density from 2750 to 2620 kgm^{-3} as the
 235 window size increases from $300 \text{ km} \times 600 \text{ km}$ to $900 \text{ km} \times 1200 \text{ km}$ (Table 1)

236 is consistent with the inclusion within the windows of a greater proportion of
237 the GSM flanks and rift basins, where the presence of lower density sedimentary
238 rocks is expected. The range of densities obtained is consistent with the averaging
239 of metamorphic basement or igneous rocks in the GSM (2700–2800 kgm⁻³) and
240 lower density sedimentary rocks in the surrounding basins (2400–2600 kgm⁻³).

241 The coherence between the FAA and topography is high over a wide range of
242 wavenumbers, particularly for the smaller windows (Fig. 3). This means there is
243 a large wavenumber band over which to confidently fit a regression line and de-
244 termine topographic density. The error associated with the admittance translates
245 as errors in ρ_c and d of ± 100 kgm⁻³ and ± 0.2 km, respectively. The observed
246 mean depth from the geodetic datum to the ice-bed interface, d , for each window
247 is within error of the mean depth recovered from spectral analysis of the uncom-
248 pensated topography.

249 The best-fitting T_e derived from the free-air admittance remains constant at
250 0–1 km for each window, which is illustrated by the computed isostatic response
251 functions (Fig. 4). Bouguer coherences indicate a slightly higher T_e of 5–14 km,
252 with a decrease in T_e as the window size is increased (Table 1; Supplementary Fig.
253 2). These anomalously low T_e estimates are in contrast with previous estimates
254 that reported instead high T_e values of ca. 70 km beneath the range and lower T_e
255 of ca. 30 km beneath the EARS inferred to surround the GSM (Ferraccioli et al.,
256 2011). These discrepancies in T_e estimates are discussed in section 5.1.

257 4.2. *Amount of Erosion and Flexure*

258 The estimated amount of eroded material in the valleys of the GSM is up to
259 1.2 km (Fig. 5c). In the Lambert Rift, 1.5–2 km of erosion is estimated, which
260 is consistent with independent estimates from ice sheet erosion models (Jamieson

et al., 2010). Detrital thermochronology shows evidence for 2–3 km of localised erosion by the Lambert Glacier since the Early Oligocene (Tochilin et al., 2012; Thomson et al., 2013).

The flexure was first computed by solving Eq (11) using an FFT method (Watts, 2001) for a continuous elastic plate with a uniform T_e of 5 km, a value consistent with the results of free-air admittance and Bouguer coherence modelling in this study. The flexural uplift in the central GSM is 500–700 m, and increases to up to 1400 m in the eastern Lambert Rift (Fig. 5d). Although there is short wavelength spatial variability in the uplift, reflecting the low rigidity of the lithosphere, profiles along strike of and perpendicular to the GSM show the flexural uplift is relatively consistent at 500–700 m throughout the range (Fig. 6).

The calculation was also carried out for $T_e = 10, 25$ and 50 km. This range of values encompasses the results of the spectral modelling in this study and the average for the Gamburtsev region determined by Ferraccioli et al. (2011). At lower T_e values, there is significant short wavelength spatial variability in the distribution of uplift. High T_e values dampen out the shorter wavelength responses, and the uplift is more widely distributed. In the Lambert Rift, $T_e = 5$ km permits localised uplift of almost 1.5 km. However, higher T_e values significantly reduce the amount of flexure; the uplift is only 700 m at $T_e = 50$ km. In the GSM, while the pattern of flexure is sensitive to T_e , the magnitude is relatively insensitive; the average uplift only decreases from 560 m ($T_e = 5$ km) to 460 m ($T_e = 50$ km) (Fig. 7; Table 2). Subtracting the flexure from the peak accordance surface gives the pre-incision topography. The pre-incision topography is 2–2.5 km in the GSM for each T_e scenario (Fig. 7).

Because East Antarctica was recently interpreted as a mosaic of distinct provinces

that came together during orogenic events (Ferraccioli et al., 2011), significant spatial variations in T_e might be expected. An alternative approach to T_e estimation was adopted by Ferraccioli et al. (2011) who used a 3D inversion based on the spatial convolution of surface and buried loads to determine a spatially variable T_e estimate for East Antarctica at 20 km horizontal resolution (Supplementary Fig. 3). The inversion incorporated bedrock topography, constraints on crustal thickness from seismic receiver function data (Hansen et al., 2010), and the extent of a dense lower crustal body proposed to reconcile misfits between observed and modelled Bouguer anomalies. Ferraccioli et al. (2011) calculate a T_e of ca. 70 km beneath the range, and ca. 30 km in the surrounding EARS.

A model incorporating an elastic plate of spatially variable thickness, using a 3D centred finite-difference technique to solve the general flexure equation, was employed to calculate the amount of flexure for this T_e scenario. In the Lambert Rift, the solution is similar to the case of uniform $T_e = 25$ km. In the GSM, where the average T_e is closer to 70 km, the amount of flexure is 400–500 m (Fig. 7).

5. Discussion

5.1. *Effective Elastic Thickness of the Gamburtsev Lithosphere*

Gravitational admittance modelling suggests that the Gamburtsev lithosphere is characterised by low T_e values. The best-fitting T_e for the free-air admittance is 0–1 km across all grid windows, and rises only slightly to 5–14 km for the Bouguer coherence. McKenzie et al. (2015) calculated the free-air admittance between Bedmap2 bedrock topography (Fretwell et al., 2013) and GOCE gravity data and determined a best-fitting average T_e for East Antarctica of 21 km. However, T_e estimation based on a spatial convolution approach suggests that the

310 Gamburtsev lithosphere is characterised by higher T_e values of ca. 70 km (Ferrac-
311 cioli et al., 2011).

312 One reason for such a discrepancy may be the role of buried/internal loads.
313 Bouguer coherence modelling indicates that the ratio of buried loading to surface
314 loading in the Gamburtsev region is approximately one (Supplementary Fig. 2).
315 Negatively buoyant loads within or at the base of the lithosphere increase the cur-
316 vature of the plate. Fitting the observed admittance with models that only incor-
317 porate surface loading will therefore cause T_e to be underestimated. In addition,
318 the windowing method may underestimate T_e because if high rigidity terranes
319 exist within the window, but are relatively localised and surrounded by low rigid-
320 ity lithosphere, the region will give the appearance of being in local, rather than
321 regional, isostatic equilibrium.

322 However, the low T_e values are borne out across all window sizes, and even
323 when buried loads are incorporated in the models for the Bouguer coherence
324 (Forsyth, 1985), relatively low T_e values (<15 km) are recovered. In addition,
325 seismic receiver function data indicate crustal thicknesses in excess of 48 km and
326 up to 57 km below the GSM (Hansen et al., 2010; Heeszel et al., 2013). Such
327 high crustal thicknesses are consistent with the long wavelength topography of the
328 GSM being dominated by Airy isostasy. At long wavelengths (>500 km), IRFs
329 (Fig. 4) appear to deviate away from elastic plate flexure curves towards finite
330 positive values of 0.3–0.5. A long wavelength correlation between gravity anoma-
331 lies and topography is unlikely to be associated with plate flexure, but has been
332 attributed to dynamic processes occurring in the Earth’s mantle (Panasyuk and
333 Hager, 2000). This may indicate that the long-wavelength elevated East Antarctic
334 plateau is - in part - dynamically supported by the convecting mantle (O’Donnell

335 and Nyblade, 2014). However, further modelling work is needed in order to better
336 understand the thermotectonic history and architecture of the Gamburtsev litho-
337 sphere, and how they link to the effective elastic thickness.

338 The inverse spectral methods, in particular the Bouguer coherence, have been
339 widely used to estimate T_e in the continents, where estimates vary from a few
340 km to over one hundred km. It has been suggested that such methods tend to
341 overestimate T_e due to the effects of erosion, which preferentially removes the
342 short wavelength components of topography (McKenzie and Fairhead, 1997). In
343 the Gamburtsevs, cold-based ice has protected the topography from erosion since
344 shortly after 34 Ma (Creyts et al., 2014; Rose et al., 2013). In addition, the free-air
345 coherences calculated in this study are among the highest ever reported for conti-
346 nental interiors (Fig. 3). Such high coherences are reminiscent of those observed
347 in the oceans (e.g. Watts, 1978), and are the result of negligible erosion rates in
348 the GSM during the last ten million years or more. Subglacial Antarctica offers
349 a previously unrecognised opportunity to evaluate the use of the inverse spectral
350 method for T_e estimation, particularly where non-erosive ice has preserved short
351 wavelength features of topography and high resolution gravity data exist.

352 5.2. *Palaeoclimate and Timing of Valley Incision*

353 The flexural response to valley incision accounts for, on average, 400–600 m
354 (17–25%) of the GSM elevation. Flexure rarely exceeds 25% of the peak elevation
355 in temperate climates (Gilchrist et al., 1994; Montgomery, 1994). This suggests
356 that the processes of valley incision in the GSM occurred in a more temperate
357 climate, and the landscape has remained unmodified since shortly after 34 Ma
358 (Creyts et al., 2014; Rose et al., 2013). Given that flexure accounts for only 17–
359 25% of the elevation of the GSM, there is a need for tectonic trigger(s) and/or

360 dynamic topography to explain the high pre-incision ancestral elevation of the
361 GSM.

362 Flexure calculations suggest that glacial incision drove 50–80% on the flanks
363 of the Lambert Rift, depending on the assumed T_e . However it is likely that the
364 Lambert Rift contains sediments of Permian age with densities lower than the
365 value of 2670 kgm^{-3} assumed in the flexure calculations, causing the amount of
366 flexure to be overestimated (Ferraccioli et al., 2011). Using a density of 2350
367 kgm^{-3} for the eroded material reduces the flexural uplift to 40–70% of the total
368 elevation. It is also likely that the difference between the peak accordance sur-
369 face and the bedrock topography is attributable not solely due to glacial erosion
370 in the Lambert Rift, but also to tectonic subsidence. This means that 40–70%
371 is an upper bound on the contribution of flexure to flank uplift. However, the
372 calculated 1.2 km of uplift in the Lambert Rift is consistent with the present el-
373 evation of Oligocene–Miocene Pagodroma Group glaciomarine sediments on the
374 Fisher Massif, now up to 1.2–1.5 km above sea-level (Hambrey and McKelvey,
375 2000; Hambrey et al., 2007). Older sediments are found at progressively higher
376 elevations, suggesting that uplift was contemporaneous with deposition. This re-
377 sult implies that a significant amount of post-Eocene uplift on the flanks of the
378 Lambert Rift can be attributed to the isostatic response to intense selective linear
379 erosion by a dynamic Lambert Glacier.

380 The flexure calculations presented in this study lend support to the hypothesis
381 that significant (2–2.5 km) topography existed in the Gamburtsev region prior to
382 34 Ma and that the mountains were a key inception point for the development
383 of the EAIS at 34 Ma (Rose et al., 2013). Geomorphometric analysis indicates
384 that an inherited fluvial landscape, which was subsequently modified by glacial

erosion, exists within the Gamburtsevs (Rose et al., 2013). One might ask for how long did this fluvial landscape exist prior to glaciation? Dating the age of the fluvial landscape of the GSM would also constrain the timing of primary uplift.

The evolution of the fluvial landscape of the GSM was simulated using a numerical model that solves the stream power equation for fluvial advection and diffusion in a temperate climate (Braun and Willett, 2013) and incorporates the ongoing isostatic response to valley incision (Appendix A). The model incorporated low long-term erosion rates as determined from detrital thermochronology (Cox et al., 2010). These erosion rates (0.01–0.02 km/Myr over the last 250 Ma) are very likely to be a minimum estimate, since they are at the lower end of the range of values derived from cosmogenic nuclide dating in Cenozoic mountain ranges (Matmon et al., 2009). It is also difficult to reconcile such low long-term erosion rates with the observation of coal beds of Permian age (Holdgate et al., 2005) and spores and pollen of palm trees of Eocene age (Pross et al., 2012) in East Antarctica, both of which indicate near-tropical climates. By assuming an erosion rate of 0.01 km/Myr, the modelled landscape age is very likely an absolute maximum.

After 50 Myr, the pattern of incision begins to resemble that of the present-day landscape. By 200 Myr, the modelled landscape closely resembles the observed landscape in terms of relief, amount of incision and position of drainage divides and major valleys (Fig. 8). After 200 Myr, the regional elevation is lowered below what is observed today. Assuming that no fluvial incision has occurred since 30 Ma, the maximum age of the preserved fluvial landscape in the GSM is 230 Ma. This result implies that mountain building occurred in interior East Antarctica after the inferred Grenvillian (Ferraccioli et al., 2011) and Pan-African (An et al.,

410 2015) orogenic events.

411 Thermochronological and structural observations are consistent with a phase
412 of exhumation and fault activity in East Antarctica during the Permo-Triassic (250
413 Ma) (Lisker et al., 2003; Phillips and Läufer, 2009), which is attributed to exten-
414 sion north of the GSM and the formation of the East Antarctic Rift System (EARS)
415 (Ferraccioli et al., 2011). Because 250 Ma is at the very upper limit for the age of
416 the fluvial landscape, Permo-Triassic rifting as the sole mechanism for GSM up-
417 lift would necessitate anomalously low erosion rates, indicative of an arid climate,
418 since 250 Ma. Low long-term erosion rates prior to Cenozoic glaciation could be
419 attributed to the long-term maintenance of a dry continental climate in interior
420 East Antarctica, which would likely necessitate large basins such as the Wilkes
421 and Aurora lying above sea-level. The presence of surficial rocks that are particu-
422 larly resistant to erosion (such as Precambrian metamorphic basement) could also
423 be a contributing factor to low long-term erosion rates.

424 In the Transantarctic Mountains, which form the boundary between East and
425 West Antarctica, the Ross Orogen was eroded to form the Kukri Peneplain, atop
426 which Devonian–Triassic Beacon Supergroup sediments were deposited in an in-
427 tracratonic/foreland basin (Elliot et al., 2015). It would appear unlikely that an
428 older (pre-Triassic) orogen in the interior of East Antarctica could survive this pro-
429 tracted period of Palaeozoic erosion and Permo-Carboniferous Gondwana glacia-
430 tion.

431 Our results indicate that more recent tectonic/dynamic uplift is needed to ex-
432 plain the high relief and heavily incised landscape of the GSM. A phase of Cre-
433 taceous exhumation in East Antarctica is attributed to the break-up of East Gond-
434 wana at 130–100 Ma (Lisker et al., 2003; Phillips and Läufer, 2009). This phase

435 of denudation may have been related to transtensional reactivation of the Permo-
436 Triassic EARS (Ferraccioli et al., 2011), although recent interpretations of detrital
437 thermochronology data appear to argue against major Cretaceous exhumation in
438 interior East Antarctica (Tochilin et al., 2012; Thomson et al., 2013).

439 *5.3. Erosion, Climate and Isostasy*

440 The total volume of estimated erosion in the GSM and Eastern Lambert Rift
441 combined is $6.2 \times 10^5 \text{ km}^3$. Jamieson et al. (2005) estimated offshore sediment
442 volumes in Prydz Bay using seismic profiles and the boundaries of glacial and
443 fluvial facies located in ODP ocean sediment cores. They estimate the presence
444 of a minimum of $54,000 \text{ km}^3$ of glaciogenic (ca. 34–0 Ma) and $98,000 \text{ km}^3$ of
445 fluvial (ca. 118–34 Ma) sediments in Prydz Bay ($1.5 \times 10^5 \text{ km}^3$ in total); the
446 total volume may be up to 10 times this value (Jamieson et al., 2005; Wilson
447 et al., 2012). These estimates are therefore in agreement to within an order of
448 magnitude. The discrepancy may arise because of the assumption that the peaks
449 in the GSM have not been lowered, the decrease in density between bedrock and
450 sediment, and because many of the valleys in the Gamburtsevs do not flow towards
451 Prydz Bay, but rather towards the basins of the South Pole, the hinterland of the
452 Transantarctic Mountains and the EARS (Fig. 8a). Much sediment was likely
453 routed towards these interior basins. More detailed geophysical study of these
454 basins is required in order to quantify the thickness of sediment present.

455 The map of eroded material (Fig. 5c) highlights two fundamentally different
456 styles of erosion - dendritic fluvial incision overprinted by Alpine-style valley
457 glaciers in the GSM and major outlet glacier-type incision in the Lambert Rift. In
458 temperate climates, the contribution of denudational isostasy to peak elevations is
459 limited by geomorphic constraints and erosion of the peaks (Gilchrist et al., 1994;

460 Whipple et al., 1999). However, selective linear glacial erosion is optimal for cre-
461 ating maximum relief in mountain ranges; basal melting is concentrated beneath
462 the thick ice in the troughs, while neighbouring peaks remain preserved beneath
463 non-erosive cold-based ice or air (Stern et al., 2005; Jamieson et al., 2014). In
464 addition, the relatively long wavelength of incision associated with broad glacial
465 outlets such as the Lambert Glacier or the Beardmore or Byrd Glaciers in the
466 Transantarctic Mountains, compared to the shorter wavelength fluvial or Alpine
467 glacial valleys seen in the GSM, permits greater flexural rebound in response to
468 unloading. The wider wavelength of incision explains why the magnitude of flex-
469 ure induced by selective linear glacial erosion is more sensitive to T_e than that
470 caused by fluvial incision.

471 Although the peaks of the Gamburtsevs likely experienced erosion prior to
472 glaciation, they have been unmodified for most of the last 34 Ma (Creyts et al.,
473 2014). While the early ice sheets in the GSM flowed down the existing river val-
474 leys (Rose et al., 2013), the modern ice sheet flows orthogonal to (and in places
475 up) the valleys (Rignot et al., 2011). Because ice is incapable of flowing fast
476 over such rough terrain due to high coefficients of basal friction, it remains cold-
477 based and non-erosive, preserving steep topographic gradients and maintaining
478 an unmodified subglacial landscape (Jamieson et al., 2014). By contrast, and
479 despite erosion in the GSM being negligible for millions of years, intense selec-
480 tive linear erosion in the Lambert Rift likely continued from the Oligocene to the
481 Neogene beneath a dynamic Lambert Glacier, which still follows the pre-existing
482 tectonically-controlled rift valley, and drove significant isostatic uplift.

483 6. Conclusions

484 In this study, we have used a combination of bedrock topography and gravity
485 data to estimate the elastic thickness of the lithosphere and spatial distribution of
486 erosion in the Gamburtsev Subglacial Mountains, in order to make a new estimate
487 of the amount of elevation that is related to erosion processes. In addition, we
488 examined the antiquity of the inherited fluvial landscape of the Gamburtsevs using
489 a landscape evolution model. Based on the results of these methods, we conclude
490 the following:

- 491 1. Free-air admittance and Bouguer coherence modelling indicates that T_e in
492 the Gamburtsev region is anomalously low (<15 km), contrasting with the
493 findings of previous studies (Ferraccioli et al., 2011). This result may be
494 indicative of a weaker-than-expected Gamburtsev lithosphere. However,
495 while T_e does influence the pattern of flexural uplift, it does not have a major
496 influence on the magnitude. Free-air coherences are among the highest ever
497 reported for the continents, reflecting negligible erosion rates during the last
498 ten million years or more.
- 499 2. The isostatic response to valley incision accounts for 17–25% of total Gam-
500 burtsev elevation, which is typical of incision in temperate climates. The
501 pre-incision topography of the GSM was 2–2.5 km. These findings lend
502 strong independent support to the hypothesis that the mountain range ex-
503 isted prior to the Eocene–Oligocene Boundary, and provided a key site for
504 EAIS nucleation. Selective glacial erosion can account for up to 70% of to-
505 tal uplift in the Lambert Rift, reflecting a markedly different erosive regime
506 that continued throughout the Oligocene–Neogene(?).

- 507 3. Assuming low long-term erosion rates, landscape evolution models indi-
508 cate that the maximum age of the inherited fluvial landscape of the GSM is
509 230 Ma. While it is unlikely that low erosion rates have persisted in East
510 Antarctica since this time, erosion rates may have been inhibited by the ex-
511 posure of Precambrian basement and the development of an arid climate
512 since Mesozoic(?) times.
- 513 4. The interaction between climate and tectonics remains a source of uncer-
514 tainty in our understanding of intraplate mountain building. In the Transantarc-
515 tic Mountains, climate - in the form of glacial incision - plays a large role
516 in the uplift of mountain peaks. In the Gamburtsevs, approximately 80%
517 of peak elevation must be attributed to tectonic/dynamic mechanisms, the
518 nature of which remains unclear.

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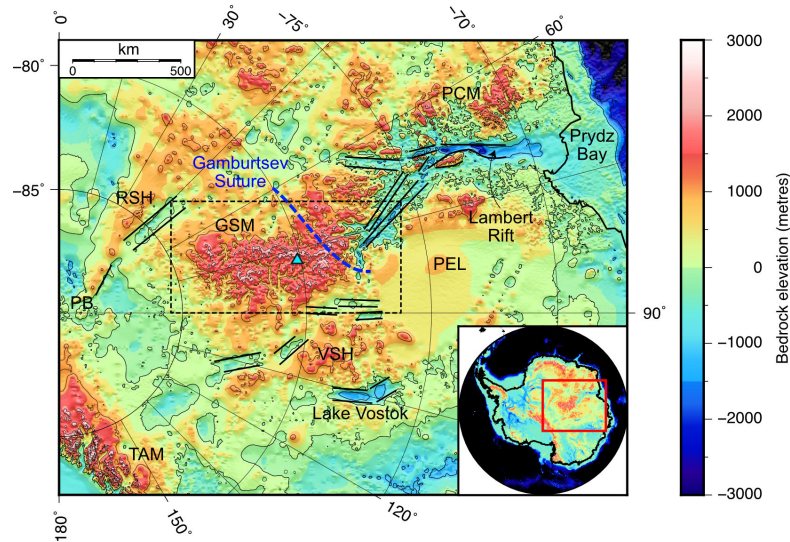


Fig. 1. Geographical and tectonic setting of the Gamburtsev Subglacial Mountains (GSM) within East Antarctica. Bedrock elevation data (above mean sea-level) are from the Bedmap2 compilation (Fretwell et al., 2013). Rift basins (bounded by black lines) comprise the recently defined East Antarctic Rift System (EARS), a proposed trigger for GSM uplift (Ferraccioli et al., 2011). The proposed location of the Gamburtsev Suture (Ferraccioli et al., 2011) is labelled with the blue dashed line. Black dashed box shows the area displayed in Figs 2 and 5. Abbreviations: PB - Polar Basins; PCM - Prince Charles Mountains; PEL - Princess Elizabeth Land; RSH - Recovery Subglacial Highlands; TAM - Transantarctic Mountains; VSH - Vostok Subglacial Highlands. Blue triangle marks Dome A. True scale at 71°S. Inset shows the main study area (red box) within Antarctica. Much of East Antarctica is characterised by an elevated topographic plateau ~1 km above sea-level.

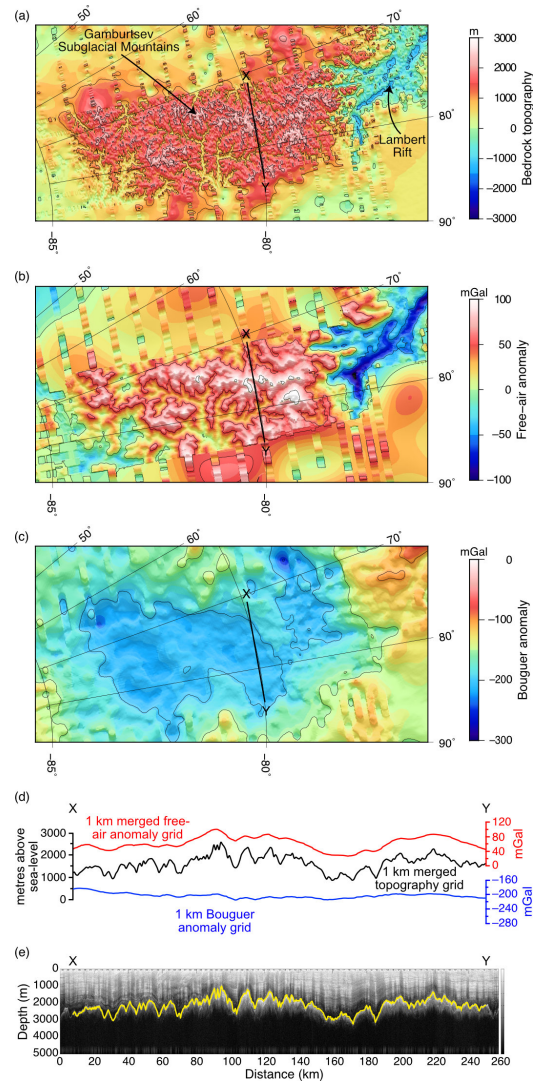


Fig. 2. Bedrock topography and gravity grids for the Gamburtsev region. (a) Bedrock topography. (b) Free-air anomaly. The AGAP free-air gravity data bear a strong resemblance to the bedrock topography, with well-defined, coherent valleys and ridges. (c) Bouguer anomaly. (d) Profile X–Y through the topography and gravity grids, illustrating the strong coherence between the FAA and topography. (e) Radar echogram for the flight line corresponding to profile X–Y. Overlain (yellow) is the bed pick, which was used along with many others to generate the bedrock topography grid.

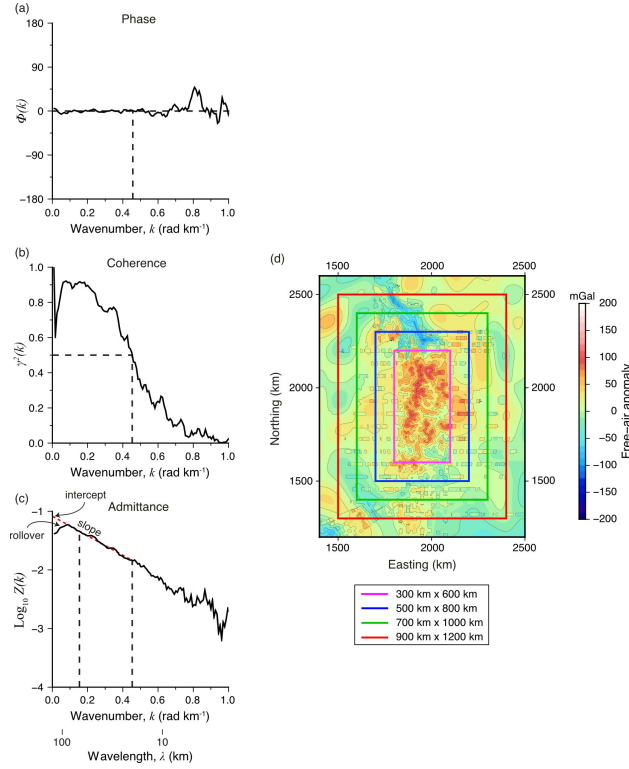


Fig. 3. Spectral parameters for the GSM. (a) Phase, $\phi(k)$. The phase remains close to zero until $k > 0.45 \text{ rad km}^{-1}$, whereafter high amplitude, short wavelength noise can be seen. (b) Coherence, $\gamma^2(k)$. The coherence is relatively high ($\gamma^2 > 0.5$) for wavenumbers less than 0.5 rad km^{-1} . At shorter wavelengths, noise causes the coherence to drop. (c) Logarithm of the admittance, $\log_{10} Z(k)$. A straight line (red dashed) was fitted (by least squares regression) to the section of the curve corresponding to uncompensated topography ($0.15 \leq k \leq k(\gamma^2 = 0.5) \text{ rad km}^{-1}$). The slope and y-intercept of this line were used to estimate the mean distance between the 4,600 m geodetic datum and the bedrock topography (d) and the mean density of the topography (ρ_c) within each grid window. (d) Calculation windows in the FAA grid projected into a local Lambert conformal conic projection. The admittance was calculated for a series of four windows of increasing size centred on the GSM. The spectral parameters plotted in (a), (b) and (c) correspond to the $300 \text{ km} \times 600 \text{ km}$ window.

Table 1. Results of the free-air admittance and Bouguer coherence modelling for each analysis window. ρ_c and d were determined by linear regression of the free-air admittance data (in the wavenumber interval ‘ k range’). The values of free-air and Bouguer T_e are those that minimised the RMS misfit between the calculated and modelled free-air admittance and Bouguer coherence, respectively.

Analysis window (km)	k range (radkm ⁻¹)	ρ (kgm ⁻³)	d (km)	Free-air T_e	Bouguer T_e
300 × 600	0.156–0.452	2750	3.60	1	14
500 × 800	0.156–0.400	2660	3.51	0	13
700 × 1000	0.156–0.382	2620	3.44	0	7
900 × 1200	0.156–0.369	2620	3.78	0	5

Table 2. Results of the erosional unloading calculations for the variety of T_e scenarios tested. The quoted uplift values are averages for the central Gamburtsev (GSM) and Lambert Rift (LR) regions.

T_e scenario	GSM uplift (m)	LR uplift (m)
Uniform 5 km	560	1250
Uniform 10 km	520	1050
Uniform 25 km	500	900
Uniform 50 km	460	700
Spatially variable	420	920

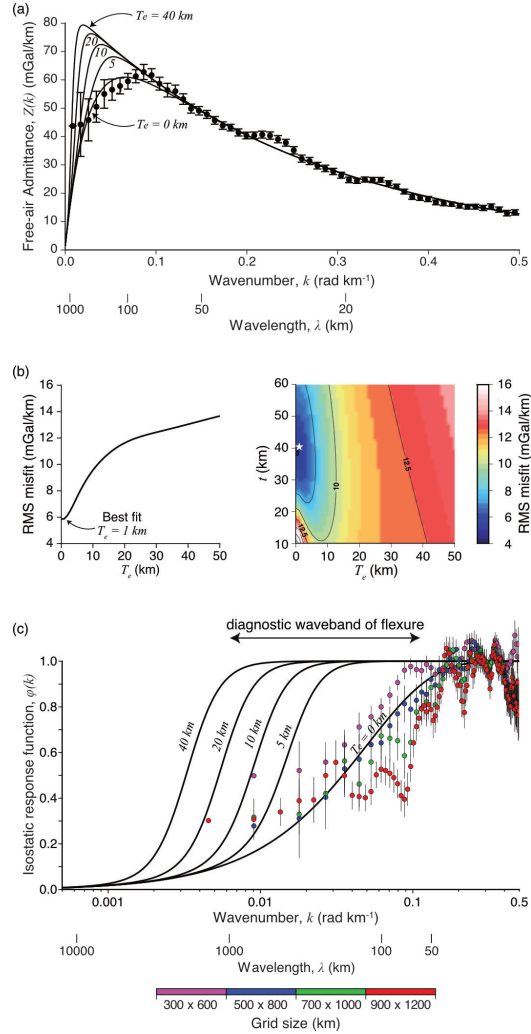


Fig. 4. Free-air admittance modelling results. (a) Free-air gravitational admittance. Black circles with standard error bars represent the calculated admittance $\pm 1\sigma$ for the 300 km \times 600 km window (Fig. 3d). The black lines are model admittance curves for an elastic plate with a crustal thickness of 40 km for varying T_e . (b) RMS misfit between calculated and modelled admittance as a function of t and T_e . Left: the value of T_e that minimises the RMS misfit is 1 km. Right: when RMS misfit is gridded as a function of t and T_e , the best-fitting value of T_e is 1 km, which occurs at $t = 41 \pm 10$ km. (c) Isostatic response functions (IRFs) for all grid windows. Coloured circles with standard error bars represent observed IRFs; solid lines are elastic plate model IRFs. The IRFs are all best-fit by a T_e of 0–5 km. At long wavelengths (> 500 km), the IRFs deviate from the model curves and tend towards finite values of 0.3–0.5. A long wavelength correlation between topography and gravity may indicate a role of mantle dynamics.

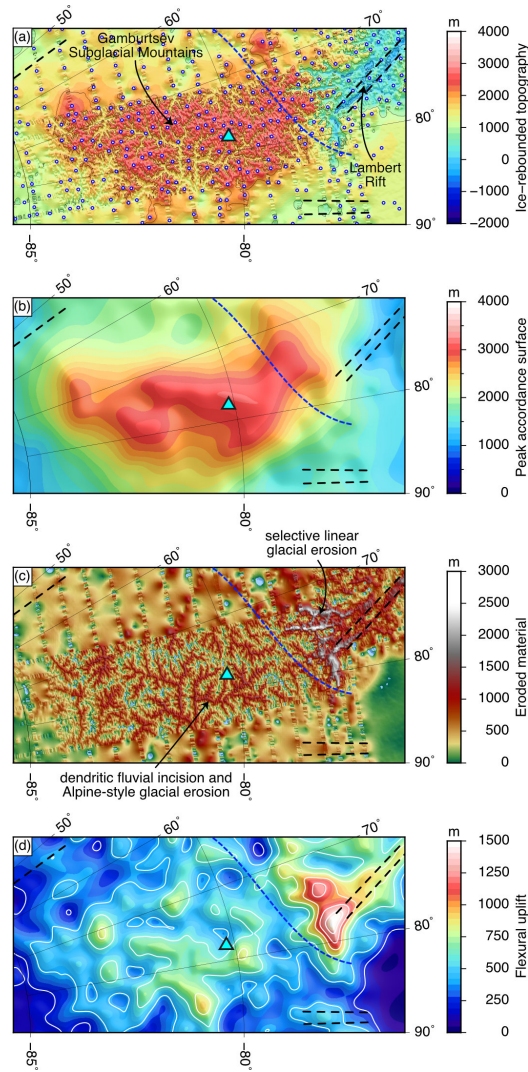


Fig. 5. Erosion-driven uplift in the Gamburtsevs. (a) Subglacial topography adjusted for ice loading assuming a continuous elastic plate model with $T_e = 5$ km. White circles with blue outlines - peaks identified using a spatial filter. (b) A surface was interpolated between the peaks and filtered with a 100 km Gaussian filter to produce a smoothed peak accordance ('cap') surface. (c) Eroded material. Calculated by subtracting the rebounded topography from the peak accordance surface. Most of the eroded material has been removed from the dendritic network of fluvial/glacial valleys in the GSM and the broad outlet glacial scours of the eastern Lambert Rift. (d) Flexural uplift. A continuous elastic plate model with $T_e = 5$ km was used. Contour intervals are 200 m. Black dashed lines - rifts of the EARS (Ferraccioli et al., 2011); blue dashed line - Gamburtsev Suture (Ferraccioli et al., 2011); Blue triangle - Dome A.

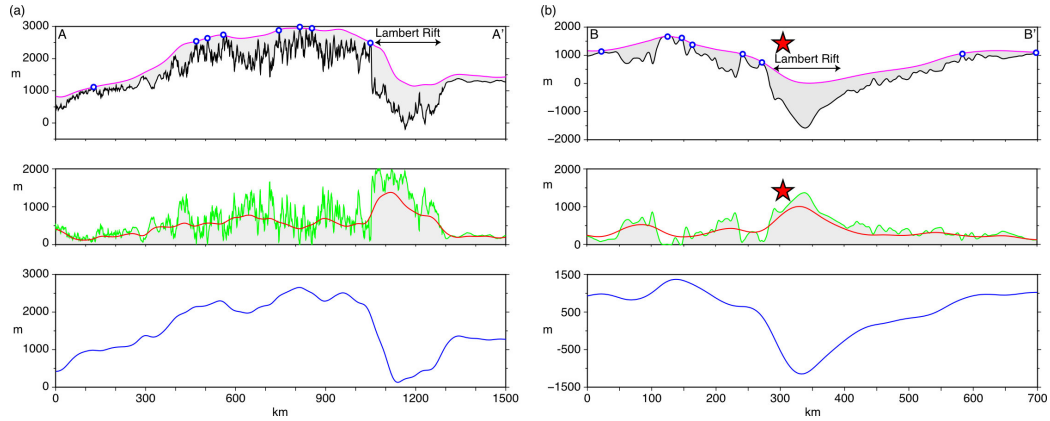


Fig. 6. Profiles (a) through the Gamburtsevs and eastern Lambert Rift from the South Pole basins to Princess Elizabeth Land (A–A') and (b) perpendicular to the strike of Lambert Rift (B–B'). Profile locations are shown in Fig. 7. Upper panel: rebounded (ice-free) topography (black line) and peak accordance surface (purple line). Circles denote peaks used to interpolate the accordance surface. The shaded region represents the eroded material. Middle panel: Eroded material (green line and shaded region) and the isostatic rebound due to the removal of the eroded material for a uniform elastic plate of $T_e = 5$ km (red line). Flexure throughout the main Gamburtsev mountain range is 500–700 m, but rises to 1.5 km in the Lambert Rift. In (b), the red star represents the present-day elevation of Oligocene–Miocene glaciomarine sediments currently exposed on Fisher Massif that were formed at sea-level. Flexure can account for over 50% of the post-Eocene uplift of these sediments. Lower panel: pre-incision topography (blue line) calculated by subtracting the flexural uplift from the peak accordance surface. This surface represents the ancestral topography that cannot be accounted for by erosion and flexure, and instead reflects primary tectonic/dynamic uplift of the Gamburtsevs and subsidence of the Lambert Rift.

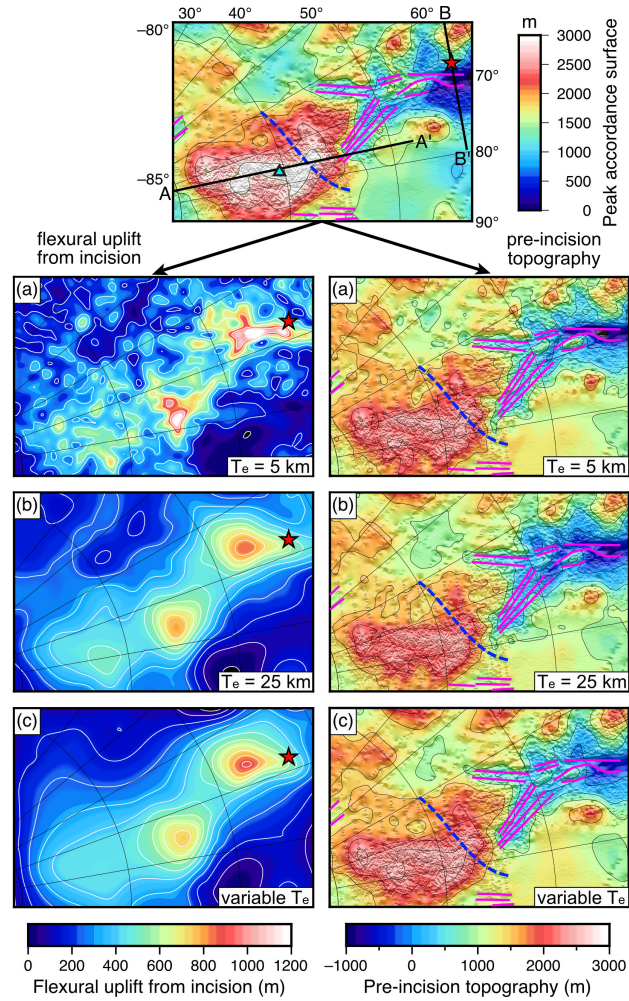


Fig. 7. Sensitivity of flexural uplift to T_e . Upper panel: peak accordance surface (cf. Fig. 5b). Purple lines - EARS; blue dashed line - Gamburtsev Suture; blue triangle - Dome A; red star - Fisher Massif. GSM uplift is partitioned into a flexural and pre-incision component for three T_e scenarios. (a) Continuous elastic plate model with uniform $T_e = 5$ km (flexural uplift is the same as shown in Fig. 5d). (b) Continuous elastic plate model with uniform $T_e = 25$ km. (c) Continuous elastic plate model with variable T_e ; ca. 70 km beneath the GSM and ca. 30 km in the surrounding EARS. Fisher Massif glaciomarine sediments indicate up to 1.5 km of post-Eocene uplift (Hambrey and McKelvey, 2000). The pre-incision topography of the GSM is 2–2.5 km for every scenario.

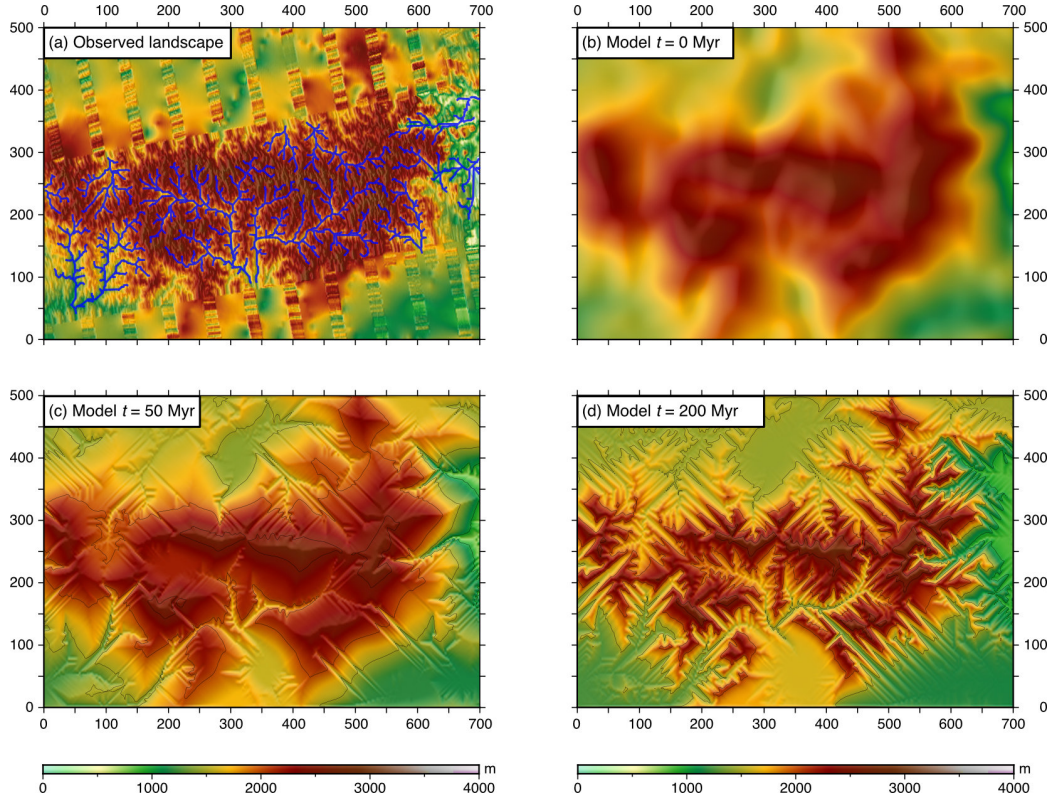


Fig. 8. Modelling the fluvial landscape evolution of the GSM. (a) Present-day Gamburtsev topography corrected for ice loading. Fluvial drainage networks (Rose et al., 2013) are shown in blue. (b) Initial topography for the landscape evolution model ($t = 0$) is the pre-incision topography calculated for $T_e = 5$ km. (c) Modelled fluvial landscape after $t = 50$ Myr. (d) Modelled fluvial landscape after $t = 200$ Myr.

531 **Appendix A. Landscape Evolution Model**

532 The landscape evolution model solves the combined fluvial incision-hillslope
533 diffusion equation.

$$\frac{\partial h}{\partial t} = U - KA^m \left(\frac{\partial h}{\partial x} \right)^n + \kappa \nabla^2 h \quad (\text{A.1})$$

534 Physically, this non-linear partial differential equation describes the advection
535 and diffusion of topography, $h(x, y, t)$, by river systems.

- 536 • The advection term, $-KA^m \left(\frac{\partial h}{\partial x} \right)^n$, is a power law function of the local drainage
537 area, A , and stream gradient, $\frac{\partial h}{\partial x}$, the quantities that control the rate of bedrock
538 channel erosion (Whipple and Tucker, 1999). K is a dimensional coefficient
539 of erosion, and depends on the erodibility of the bedrock and the amount of
540 rain that falls (which is assumed to be constant throughout the domain). m
541 and n are positive power law exponents, commonly taken as $\frac{1}{3}$ and 1, re-
542 spectively (Whipple and Tucker, 1999).
- 543 • The hillslope diffusion term, $\kappa \nabla^2 h$, takes the form of a typical 2D diffusion
544 equation; the ‘erosional diffusivity’ is given by κ .
- 545 • The uplift term, U , incorporates the ongoing isostatic adjustment to the re-
546 moval of mass by river systems.

547 Water is rained onto a regularly spaced topographic surface. Each grid node is
548 surrounded by 8 neighbours. A D8 streamflow algorithm distributes water to the
549 neighbouring grid square with the lowest elevation. Rivers are permitted to flow
550 out of the domain through any of the boundaries. Eq (A.1) is solved numerically
551 at a series of timesteps. A series of assumptions were made in assigning values to
552 the free parameters in Eq (A.1).

553 The GSM topography is very rugged (Fig. 2), which suggests fluvial advection
 554 dominates over diffusion. Preliminary model runs showed that the final landscape
 555 is insensitive to κ ; the diffusion term was neglected. Eq (A.1) is reduced to

$$\frac{\partial h}{\partial t} = U - KA^{\frac{1}{3}} \frac{\partial h}{\partial x} \quad (\text{A.2})$$

556 The typical value of the advection constant, K , in modern mountain ranges, such
 557 as the European Alps, is $10^{-5} \text{ m}^{\frac{1}{3}} \text{ yr}^{-1}$ (Whipple and Tucker, 1999). Average
 558 Pliocene–Pleistocene erosion rates in the Alps are 0.1–0.5 km/Myr (Champagnac
 559 et al., 2007); detrital AFT thermochronology from Prydz Bay sediments suggests
 560 minimum long-term erosion rates of 0.01 km/Myr in East Antarctica (Cox et al.,
 561 2010). Assuming that the ratio between K and measured erosion rates is the same
 562 in the Alps and the Gamburtsevs (which exhibit strikingly similar relief and ge-
 563 omorphology, implying that the erosional regimes were similar), the minimum
 564 value of K for the GSM is $2 \times 10^{-7} \text{ m}^{\frac{1}{3}} \text{ yr}^{-1}$.

565 The initial topography, $h(x, y, 0)$, was the pre-incision topography grid derived
 566 in this study (for $T_e = 5 \text{ km}$), which represents the cumulative tectonic/dynamic
 567 uplift in the absence of incision. The grid was resampled to a resolution of 2 km
 568 to ease the computational demand.

569 The model uses a numerical integration to discretise a continuous process; the
 570 upstream drainage area and local slope are calculated at each timestep, as is the
 571 uplift due to erosional unloading. The uplift term, $U(x, y, t)$, was calculated using
 572 a viscoelastic plate model, which applies the correspondence principle (Brotchie
 573 and Silvester, 1969) to derive the viscoelastic flexure ($W(k, t)$) from the initial
 574 ($W(k, 0)$) and final ($W(k, \infty)$) elastic response.

$$W(k, t) = W(k, 0) e^{-t/\tau} + [1 - e^{-t/\tau}] [W(k, \infty) - W(k, 0)] \quad (\text{A.3})$$

575 The viscoelastic parameters assigned to the model were a Maxwell relaxation
576 time, τ of 0.01 Myr (corresponding to an effective viscosity of 10^{22} Pa s), an
577 initial T_e of 90 km (the typical seismic thickness of the lithosphere), and a final T_e
578 of 5 km.

579 **References**

- 580 An, M., Wiens, D. A., Zhao, Y., Feng, M., Nyblade, A. A., Kanao, M., Li, Y.,
581 Maggi, A., L  v  que, J.-J., 2015. S-velocity model and inferred Moho topogra-
582 phy beneath the Antarctic Plate from Rayleigh waves. *Journal of Geophysical*
583 *Research: Solid Earth* 120 (1), 359–383.
- 584 Bell, R. E., Ferraccioli, F., Creyts, T. T., Braaten, D., Corr, H., Das, I., Damaske,
585 D., Frearson, N., Jordan, T., Rose, K., Studinger, M., Wolovick, M., 2011.
586 Widespread persistent thickening of the East Antarctic ice sheet by freezing
587 from the base. *Science* 331 (6024), 1592–1595.
- 588 Bo, S., Siegert, M. J., Mudd, S. M., Sugden, D., Fujita, S., Xiangbin, C., Yunyun,
589 J., Xueyuan, T., Yuansheng, L., 2009. The Gamburtsev mountains and the ori-
590 gin and early evolution of the Antarctic Ice Sheet. *Nature* 459 (7247), 690–693.
- 591 Braun, J., Willett, S. D., 2013. A very efficient $O(n)$, implicit and parallel method
592 to solve the stream power equation governing fluvial incision and landscape
593 evolution. *Geomorphology* 180-181, 170–179.
- 594 Brothie, J. F., Silvester, R., 1969. On crustal flexure. *Journal of Geophysical*
595 *Research* 74 (22), 5240–5252.
- 596 Champagnac, J. D., Molnar, P., Anderson, R. S., Sue, C., Delacou, B., 2007. Qua-
597 ternary erosion-induced isostatic rebound in the western Alps. *Geology* 35 (3),
598 195–198.
- 599 Cox, S. E., Thomson, S. N., Reiners, P. W., Hemming, S. R., van de Flierdt, T.,
600 2010. Extremely low long-term erosion rates around the Gamburtsev Mountains
601 in interior East Antarctica. *Geophysical Research Letters* 37 (22), L22307.

602 Creyts, T. T., Ferraccioli, F., Bell, R. E., Wolovick, M., Corr, H., Rose, K. C.,
603 Frearson, N., Damaske, D., Jordan, T., Braaten, D., Finn, C., 2014. Freezing of
604 ridges and water networks preserves the Gamburtsev Subglacial Mountains for
605 millions of years. *Geophysical Research Letters* 41, 8114–8122.

606 DeConto, R. M., Pollard, D., 2003. Rapid Cenozoic glaciation of Antarctica in-
607 duced by declining atmospheric CO₂. *Nature* 421 (6920), 245–249.

608 Elliot, D. H., Fanning, C. M., Hulett, S. R. W., 2015. Age provinces in the Antarc-
609 tic craton: Evidence from detrital zircons in Permian strata from the Beardmore
610 Glacier region, Antarctica. *Gondwana Research* 28 (1), 152–164.

611 Ferraccioli, F., Finn, C. A., Jordan, T. A., Bell, R. E., Anderson, L. M., Damaske,
612 D., 2011. East Antarctic rifting triggers uplift of the Gamburtsev Mountains.
613 *Nature* 479 (7373), 388–392.

614 Forsyth, D. W., 1985. Subsurface loading and estimates of the flexural rigidity
615 of continental lithosphere. *Journal of Geophysical Research* 90 (B14), 12623–
616 12632.

617 Fretwell, P., Pritchard, H. D., Vaughan, D. G., Bamber, J. L., Barrand, N. E., Bell,
618 R., Bianchi, C., Bingham, R. G., Blankenship, D. D., Casassa, G., Catania, G.,
619 Callens, D., Conway, H., Cook, A. J., Corr, H. F. J., Damaske, D., Damm, V.,
620 Ferraccioli, F., Forsberg, R., Fujita, S., Gim, Y., Gogineni, P., Griggs, J. A.,
621 Hindmarsh, R. C. A., Holmlund, P., Holt, J. W., Jacobel, R. W., Jenkins, A.,
622 Jokat, W., Jordan, T., King, E. C., Kohler, J., Krabill, W., Riger-Kusk, M.,
623 Langle, K. A., Leitchenkov, G., Leuschen, C., Luyendyk, B. P., Matsuoka, K.,
624 Mougnot, J., Nitsche, F. O., Nogi, Y., Nost, O. A., Popov, S. V., Rignot, E.,

- 625 Rippin, D. M., Rivera, A., Roberts, J., Ross, N., Siegert, M. J., Smith, A. M.,
626 Steinhage, D., Studinger, M., Sun, B., Tinto, B. K., Welch, B. C., Wilson, D.,
627 Young, D. A., Xiangbin, C., Zirizzotti, A., feb 2013. Bedmap2: improved ice
628 bed, surface and thickness datasets for Antarctica. *The Cryosphere* 7 (1), 375–
629 393.
- 630 Gilchrist, A. R., Summerfield, M. A., Cockburn, H. A. P., 1994. Landscape dis-
631 section, isostatic uplift, and the morphologic development of orogens. *Geology*
632 22, 963–966.
- 633 Hambrey, M. J., Glasser, N., McKelvey, B., Sugden, D., Fink, D., 2007. Ceno-
634 zoic landscape evolution of an East Antarctic oasis (Radok Lake area, northern
635 Prince Charles Mountains), and its implications for the glacial and climatic his-
636 tory of Antarctica. *Quaternary Science Reviews* 26 (5-6), 598–626.
- 637 Hambrey, M. J., McKelvey, B., 2000. Major Neogene fluctuations of the East
638 Antarctic ice sheet: Stratigraphic evidence from the Lambert Glacier region.
639 *Geology* 28 (10), 887–890.
- 640 Hansen, S. E., Nyblade, A. A., Heeszel, D. S., Wiens, D. A., Shore, P., Kanao,
641 M., 2010. Crustal structure of the Gamburtsev Mountains, East Antarctica, from
642 S-wave receiver functions and Rayleigh wave phase velocities. *Earth and Plan-
643 etary Science Letters* 300 (3-4), 395–401.
- 644 Heeszel, D. S., Wiens, D. A., Nyblade, A. A., Hansen, S. E., Kanao, M., An,
645 M., Zhao, Y., 2013. Rayleigh wave constraints on the structure and tectonic
646 history of the Gamburtsev Subglacial Mountains, East Antarctica. *Journal of
647 Geophysical Research: Solid Earth* 118 (5), 2138–2153.

- 648 Holdgate, G. R., McLoughlin, S., Drinnan, A. N., Finkelman, R. B., Willett, J. C.,
649 Chiehowsky, L. A., 2005. Inorganic chemistry, petrography and palaeobotany
650 of Permian coals in the Prince Charles Mountains, East Antarctica. *International*
651 *Journal of Coal Geology* 63 (1-2 SPEC. ISS.), 156–177.
- 652 Holt, J. W., Richter, T. G., Kempf, S. D., Morse, D. L., Blankenship, D. D., 2006.
653 Airborne gravity over Lake Vostok and adjacent highlands of East Antarctica.
654 *Geochemistry, Geophysics, Geosystems* 7 (11), 1–15.
- 655 Jamieson, S. S. R., Hulton, N. R. J., Sugden, D. E., Payne, A. J., Taylor, J., 2005.
656 Cenozoic landscape evolution of the Lambert basin, East Antarctica: the rela-
657 tive role of rivers and ice sheets. *Global and Planetary Change* 45, 35–49.
- 658 Jamieson, S. S. R., Stokes, C. R., Ross, N., Rippin, D. M., Bingham, R. G., Wil-
659 son, D. S., Margold, M., Bentley, M. J., 2014. The glacial geomorphology of
660 the Antarctic ice sheet bed. *Antarctic Science* 26 (06), 724–741.
- 661 Jamieson, S. S. R., Sugden, D. E., Hulton, N. R., 2010. The evolution of the
662 subglacial landscape of Antarctica. *Earth and Planetary Science Letters* 293 (1-
663 2), 1–27.
- 664 Jordan, T. A., Ferraccioli, F., Corr, H., Robinson, C., Caneva, G., Armadillo,
665 E., Bozzo, E., Frearson, N., 2007. Linking the Wilkes Subglacial Basin, the
666 Transantarctic Mountains, and the Ross Sea with a New Airborne Gravity Sur-
667 vey. *Terra Antarctica Reports* 13, 37–54.
- 668 Kirby, J. F., 2014. Estimation of the effective elastic thickness of the lithosphere
669 using inverse spectral methods: The state of the art. *Tectonophysics* 631, 87–
670 116.

- 671 Lisker, F., Brown, R., Fabel, D., 2003. Denudational and thermal history along
672 a transect across the Lambert Graben, northern Prince Charles Mountains,
673 Antarctica, derived from apatite fission track thermochronology. *Tectonics*
674 22 (5), 1055.
- 675 Matmon, A., Simhai, O., Amit, R., Haviv, I., Porat, N., McDonald, E., Benedetti,
676 L., Finkel, R., 2009. Desert pavement-coated surfaces in extreme deserts
677 present the longest-lived landforms on Earth. *Bulletin of the Geological So-*
678 *ciet of America* 121 (5-6), 688–697.
- 679 McKenzie, D., 2003. Estimating T_e in the presence of internal loads. *Journal of*
680 *Geophysical Research* 108 (B9), 1–21.
- 681 McKenzie, D., Daly, M. C., Priestley, K., 2015. The lithospheric structure of
682 Pangea. *Geology* 43 (9), 783–786.
- 683 McKenzie, D., Fairhead, D., 1997. Estimates of the effective elastic thickness of
684 the continental lithosphere from Bouguer and free air gravity anomalies. *Journal*
685 *of Geophysical Research* 102 (B12), 27523–27552.
- 686 Montgomery, D. R., 1994. Valley incision and the uplift of mountain peaks. *Jour-*
687 *nal of Geophysical Research* 99 (B7), 13913–13921.
- 688 O'Donnell, J. P., Nyblade, A. A., 2014. Antarctica's hypsometry and crustal thick-
689 ness: Implications for the origin of anomalous topography in East Antarctica.
690 *Earth and Planetary Science Letters* 388, 143–155.
- 691 Panasyuk, S. V., Hager, B. H., 2000. Models of isostatic and dynamic topogra-
692 phy, geoid anomalies, and their uncertainties. *Journal of Geophysical Research*
693 105 (B12), 28199–28209.

- 694 Parker, R. L., 1972. The Rapid Calculation of Potential Anomalies. *Geophysical*
695 *Journal of the Royal Astronomical Society* 31 (4), 447–455.
- 696 Pérez-Gussinyé, M., Lowry, A. R., Watts, A. B., Velicogna, I., 2004. On the re-
697 covery of effective elastic thickness using spectral methods: Examples from
698 synthetic data and from the Fennoscandian Shield. *Journal of Geophysical Re-*
699 *search* 109 (B10), 1–20.
- 700 Phillips, G., Läufer, A. L., 2009. Brittle deformation relating to the Carbonifer-
701 ousCretaceous evolution of the Lambert Graben, East Antarctica: A precursor
702 for Cenozoic relief development in an intraplate and glaciated region. *Tectono-*
703 *physics* 471 (3-4), 216–224.
- 704 Pross, J., Contreras, L., Bijl, P. K., Greenwood, D. R., Bohaty, S. M., Schouten,
705 S., Bendle, J. A., Röhl, U., Tauxe, L., Raine, J. I., Huck, C. E., van de Flierdt,
706 T., Jamieson, S. S. R., Stickley, C. E., van de Schootbrugge, B., Escutia, C.,
707 Brinkhuis, H., 2012. Persistent near-tropical warmth on the Antarctic continent
708 during the early Eocene epoch. *Nature* 488 (7409), 73–7.
- 709 Rignot, E., Mouginot, J., Scheuchl, B., 2011. Ice Flow of the Antarctic Ice Sheet.
710 *Science* 333 (6048), 1427–1430.
- 711 Rose, K. C., Ferraccioli, F., Jamieson, S. S. R., Bell, R. E., Corr, H., Creyts,
712 T. T., Braaten, D., Jordan, T. A., Fretwell, P. T., Damaske, D., 2013. Early
713 East Antarctic Ice Sheet growth recorded in the landscape of the Gamburtsev
714 Subglacial Mountains. *Earth and Planetary Science Letters* 375, 1–12.
- 715 Stern, T. A., Baxter, A. K., Barrett, P. J., 2005. Isostatic rebound due to glacial
716 erosion within the Transantarctic Mountains. *Geology* 33 (3), 221–224.

- 717 Studinger, M., Bell, R., Frearson, N., 2008. Comparison of AIRGrav and GT-1A
718 airborne gravimeters for research applications. *Geophysics* 73 (6), 151–161.
- 719 Thomson, S. N., Reiners, P. W., Hemming, S. R., Gehrels, G. E., 2013. The con-
720 tribution of glacial erosion to shaping the hidden landscape of East Antarctica.
721 *Nature Geoscience* 6 (3), 203–207.
- 722 Tochilin, C. J., Reiners, P. W., Thomson, S. N., Gehrels, G. E., Hemming, S. R.,
723 Pierce, E. L., 2012. Erosional history of the Prydz Bay sector of East Antarc-
724 tica from detrital apatite and zircon geo- and thermochronology multidating.
725 *Geochemistry, Geophysics, Geosystems* 13 (11), 1–21.
- 726 Watts, A. B., 1978. An analysis of isostasy in the world's oceans 1. Hawaiian-
727 Emperor Seamount Chain. *Journal of Geophysical Research* 83 (B12), 5989–
728 6004.
- 729 Watts, A. B., 2001. *Isostasy and Flexure of the Lithosphere*. Cambridge University
730 Press, Cambridge.
- 731 Wessel, P., Smith, W. H. F., Scharroo, R., Luis, J., Wobbe, F., 2013. Generic
732 Mapping Tools: Improved Version Released. *Eos, Transactions American Geo-*
733 *physical Union* 94 (45), 409–410.
- 734 Whipple, K. X., Kirby, E., Brocklehurst, S. H., 1999. Geomorphic limits to
735 climate-induced increases in topographic relief. *Nature* 401, 39–43.
- 736 Whipple, K. X., Tucker, G. E., 1999. Dynamics of the stream-power river incision
737 model: Implications for height limits of mountain ranges, landscape response
738 timescales, and research needs. *Journal of Geophysical Research* 104 (B8),
739 17661–17674.

- 740 Wilson, D. S., Jamieson, S. S. R., Barrett, P. J., Leitchenkov, G., Gohl, K., Larter,
741 R. D., 2012. Antarctic topography at the Eocene-Oligocene boundary. *Palaeo-*
742 *geography, Palaeoclimatology, Palaeoecology* 335-336, 24–34.
- 743 Yi, W., Rummel, R., Gruber, T., apr 2013. Gravity field contribution analysis
744 of GOCE gravitational gradient components. *Studia Geophysica et Geodaetica*
745 57 (2), 174–202.